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Author for correspondence:
Robert J. Stern
e-mail: rjstern@utdallas.edu

The evolution of plate tectonics

Robert J. Stern

Geosciences Department, University of Texas at Dallas, Richardson, TX 75080, USA

RJS, 0000-0002-8083-4632

To understand how plate tectonics became Earth's dominant mode of convection, we need to address three related problems. (i) What was Earth's tectonic regime before the present episode of plate tectonics began? (ii) Given the preceding tectonic regime, how did plate tectonics become established? (iii) When did the present episode of plate tectonics begin? The tripartite nature of the problem complicates solving it, but, when we have all three answers, the requisite consilience will provide greater confidence than if we only focus on the long-standing question of when did plate tectonics begin? Earth probably experienced episodes of magma ocean, heat-pipe, and increasingly sluggish single lid magmatotectonism. In this effort we should consider all possible scenarios and lines of evidence. As we address these questions, we should acknowledge there were probably multiple episodes of plate tectonic and non-plate tectonic convective styles on Earth. Non-plate tectonic styles were probably dominated by 'single lid tectonics' and this evolved as Earth cooled and its lithosphere thickened. Evidence from the rock record indicates that the modern episode of plate tectonics began in Neoproterozoic time. A Neoproterozoic transition from single lid to plate tectonics also explains kimberlite ages, the Neoproterozoic climate crisis and the Neoproterozoic acceleration of evolution.

This article is part of a discussion meeting issue 'Earth dynamics and the development of plate tectonics'.

1. Introduction

Plate tectonics is the central unifying theory for geology and geophysics. The original definition of plate tectonics [1] has recently been modified to include a description of the driving force as 'A theory of global tectonics powered by subduction in which the lithosphere is divided into a mosaic of plates, which move on and sink into weaker ductile asthenosphere.

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Three types of localized plate boundaries form the interconnected global network: new oceanic plate material is created by seafloor spreading at *mid-ocean ridges*, old oceanic lithosphere sinks at *subduction zones*, and two plates slide past each other along *transform faults*. The negative buoyancy of old dense oceanic lithosphere, which sinks in subduction zones, mostly powers plate movements' [2]. Plate tectonics is a special planetary convection style that is characterized by the largely independent motion of lithospheric fragments, with new lithosphere formed at mid-ocean ridges and consumed in subduction zones. This is what is meant by 'the modern episode of plate tectonics' in the rest of the text.

We want to understand how plate tectonics became the defining convective style of our planet: the evolution of plate tectonics. To do this, we need to address three related problems. (i) What was Earth's tectonic style before the modern episode of plate tectonics began? (ii) How did plate tectonic begin? (iii) When did the modern episode of plate tectonics begin? The three-headed nature of the problem complicates arriving at a solution but, when we have all three answers, the requisite consilience will provide confidence that useful answers have been arrived at. In this effort we should keep an open mind and consider all possible scenarios and lines of evidence. For example, we should consider the possibility that there have been multiple episodes of plate tectonics in Earth history and that the transition from one tectonic/convective mode to another could have taken tens to hundreds of millions of years to accomplish.

This paper critically examines these three related topics. Although the structure of this paper as outlined above is robust, many of the conclusions are likely to be controversial because I interpret the physical constraints and the geological evidence to indicate that plate tectonics has only operated for the last quarter or so of Earth history, beginning in Neoproterozoic time. In addition, this essay also explores 'other things explained'. Because the transition to plate tectonics must have affected all Earth systems, including the biosphere and hydrosphere, answering the three central questions should also provide additional, unexpected insights into Earth history. Included among the 'other things explained' by this interpretation are the kimberlite record, Neoproterozoic climate change and acceleration of biological evolution in Neoproterozoic time. All these points are examined further below.

2. What was Earth's tectonic style before plate tectonics?

We want to understand what was Earth's tectonic and convective style before the modern episode of plate tectonics was established. We can gain a useful perspective on the range of Earth's possible tectonic and magmatic styles by considering other active large (greater than 1000 km diameter), rocky (silicate) bodies in our Solar System. In 2015, mankind completed a first examination of the 30 largest bodies orbiting the Sun, the four largest of which (Jupiter, Saturn, Uranus and Neptune) are gassy or icy giants not considered here. The other 26 largest solid bodies in the Solar System can be subdivided on a basis of density into eight rocky ($\rho > 3000\,\mathrm{kg\,m^{-3}}$, inferred to be mostly silicates) and 18 icy ($\rho < 2200\,\mathrm{kg\,m^{-3}}$) bodies [3]. For the purposes of this discussion we ignore the icy bodies because their very different rheologies and densities complicates comparisons with Earth.

Next we should identify silicate bodies that are tectonically active and focus on these for comparison with Earth.

Before doing this it is useful to consider the roles of lithosphere and asthenosphere. The concept of lithosphere has evolved considerably since it was introduced by Barrell [4] to describe the strong outer shell or lid of the Earth. Earth's lithosphere has three useful definitions: thermal, chemical and rheological/mechanical [5]. Thermal lithosphere has a conductive geothermal gradient; chemical lithosphere has compositional and isotopic characteristics that reflect prolonged isolation from well-mixed asthenosphere; and rheological (mechanical) lithosphere is defined by strength. So defined, these different lithospheres have different thicknesses; generally thermal and chemical lithosphere are subequal and much thicker than rheological lithosphere. For tectonically active planets, lithosphere is underlain by asthenosphere, which is hotter and weak enough to flow. Earth's asthenosphere lies a few hundred kilometres below the surface but we

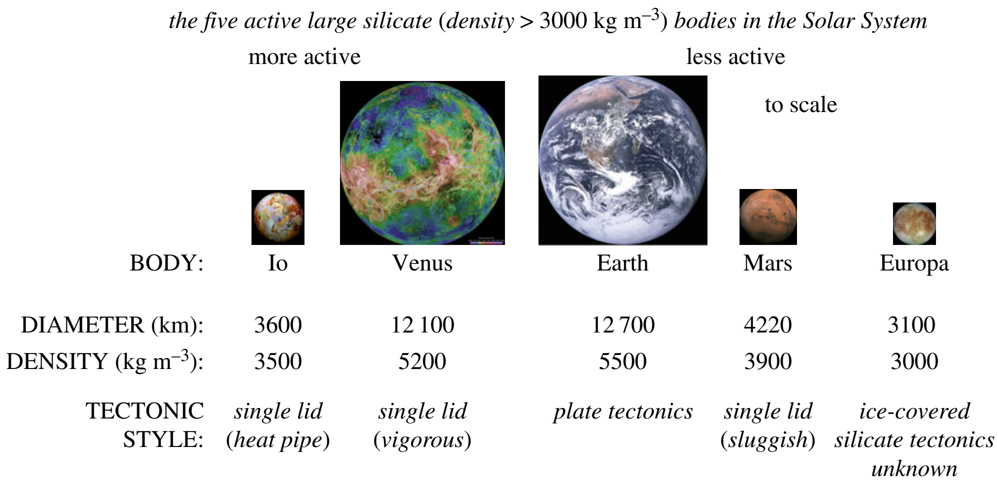

Figure 1. Images of the solid surfaces of the five large (greater than 1000 km diameter), active (TAI = 2 or greater; [3]), silicate (density = 3000 kg km^{-3} or greater) bodies. Three of the five (Io, Venus and Mars) have some variety of single lid tectonics. We do not know what is the tectonic style of the silicate interior of Europa because the planet is covered in ice. Only Earth has plate tectonics.

Table 1. Active silicate planetary bodies (planets in italics).

body	diameter (km)	mass (kg)	density (kg m^{-3})	TAI[a]	outer lithosphere	tectonic style	plumes?
Venus	12 100	4.90×10^{24}	5.28×10^{3}	3	silicate	single lid	yes
Earth	12 700	6.00×10^{24}	5.59×10^{3}	3	silicate	fragmented lid	yes
Mars	6779	6.40×10^{23}	3.92×10^{3}	2	silicate	single lid	yes
Io	3600	8.90×10^{22}	3.64×10^{3}	2	silicate	single lid	yes
Europa	3100	4.80×10^{22}	3.08×10^{3}	2	ice	fragmented (ice) lid	yes

[a]Tectonic activity index.

have few constraints on lithospheric thicknesses on other active silicate bodies. On Earth, mantle convection via plate tectonics is reflected in deformation (folding and faulting of lithosphere) and volcanism (asthenospheric melts that reach the surface) and similar features are expected on other active silicate bodies: Venus, Mars and Io. Tectonically dead silicate bodies are easily recognized by surfaces that are pockmarked by impacts, like those of Mercury and Earth's moon.

We can use the above considerations to assess whether a silicate body is active or not. The tectonic activity index (TAI) uses observed surface features to yield a three-point score summarizing a body's volcanism, faulting and density of bolide impacts. Combining density information and TAI allows us to identify the large active silicate bodies [3]. Table 1 summarizes the five active silicate bodies in our Solar System: three planets (Venus, Earth and Mars) and two satellites of Jupiter (Io and Europa) (figure 1). Europa is classified as an active body because its icy surface is faulted and has few impacts but because it is covered in ice we cannot assess whether its silicate interior is active or not. For this reason it is omitted from further consideration and focus is on the four active silicate bodies with observable surfaces (Venus, Earth, Mars and Io).

Among the four active silicate bodies listed in table 1, only Earth has plate tectonics. The other three have a tectonic style where the lithosphere is not broken into largely independent mobile plates but is a single lithospheric 'lid' that completely encloses the asthenosphere and the interior of the body. This idea was originally advanced by Solomatov & Moresi [6] for Venus, who argued that convection in its interior was dominated by upwelling plumes similar to those observed on Earth, and by cold 'drips' of lithosphere that sink into the asthenosphere. Solomatov & Moresi [6] coined the term 'stagnant lid' and used it to characterize the present Venusian tectonic regime. Venusian lithosphere is deformed along lithospheric weak zones and is accompanied by widespread magmatism (Venus-style squishy lid [7]). Similarly, Mars has a stagnant lid tectonic regime, albeit more sluggish than the Venusian tectonomagmatic regime (e.g. [8]).

The idea that large silicate bodies tend to have an all-encompassing lithospheric lid finds increasing acceptance by planetary scientists. Our understanding of this non-plate tectonic regime is evolving rapidly and we are now exploring what are the possible styles of 'lid tectonics' for large silicate bodies. We recognize three active examples (Venus, Mars and Io) but we have only glimpsed their surfaces. Names for the wide range of single lid tectonic behaviours are proliferating rapidly. For example, O'Neill & Roberts [9] refer to 'stagnant, sluggish, plutonic-squishy, or heat pipe' variants; in addition there is plume-lid tectonics [10]. 'Sagduction'—the vertical sinking of weak lithosphere—is another vigorous non-plate tectonic style [11]. Here I use the term 'single lid' to describe all types of tectonic and magmatic activity in a convecting and magmatically active silicate body where the lithosphere is not fragmented into plates. It must be stressed that we know very little about the range of single lid behaviours [12], but the proliferation of new terms for these indicates that we have begun the serious work needed to achieve this understanding.

Given that Earth may have experienced one or more episodes of single lid behaviour, how can we identify these in the rock record? We cannot do that with any confidence yet, but we should keep in mind that some transitions that have been identified in the rock record—for example, different trace element and isotopic ratios preserved in sediments and igneous rocks—might mark changes in single lid behaviour and not necessarily the beginning of plate tectonics. Part of the reason for this uncertainty is that we have an incomplete understanding of the range of single lid behaviours, largely due to the fact that we cannot yet study in detail any of the three single lid examples in our Solar System. (The Mars InSight mission (scheduled to land on Mars in November 2018) promises to make the first measurements of an active single lid silicate body interior.) We expect that mantle plume activity will be an important mode of convective upwelling in active single lid silicate bodies and that loss of lithosphere via Rayleigh–Taylor instabilities or 'drips' will be important modes of convective downwelling. Interactions between plume-derived magmas and pre-existing continental crust may yield incompatible trace element abundances that mimic those of convergent margin magmas (e.g. [13]). Styles of plume upwellings and drip downwellings will change as the body cools and lithosphere thickens. There is petrological evidence that Earth's potential temperature—approximating the temperature of the asthenosphere—has decreased by approximately 150–300°C over the past 2.5 Ga [14,15], and cooler asthenosphere implies thicker mantle lithosphere and thinner oceanic (basaltic) crust. The expected changes in the thickness of oceanic crust and mantle lithosphere over the life of a cooling planet should be manifested in lithospheric behaviour, not only in terms of whether plate tectonics or some other tectonic style exists but also in terms of changing styles of single lid behaviour. Thin mantle lithosphere at the beginning of a silicate body's evolution is likely to be associated with different types of upwellings and downwellings than those associated with thicker mantle lithosphere later in the history of the same silicate body. A cartoon of this based on the behaviour of hyperactive Io, active Venus and sluggish Mars is shown in figure 2.

The planetary perspective indicates that plate tectonics is unusual. Given that single lid tectonics is the most common convective mode of large, active silicate bodies in the Solar System, plate tectonics may also be unusual in Earth history. A tectonic style that is unusual in space may also be unusual in time. Earth-centric uniformitarianism suggests that plate tectonics has always operated, but planetary uniformitarianism yields a very different conclusion. Based on what we have learned from exploring the Solar System, it is very likely that Earth experienced one or more episodes of single lid behaviour before modern plate tectonics began.

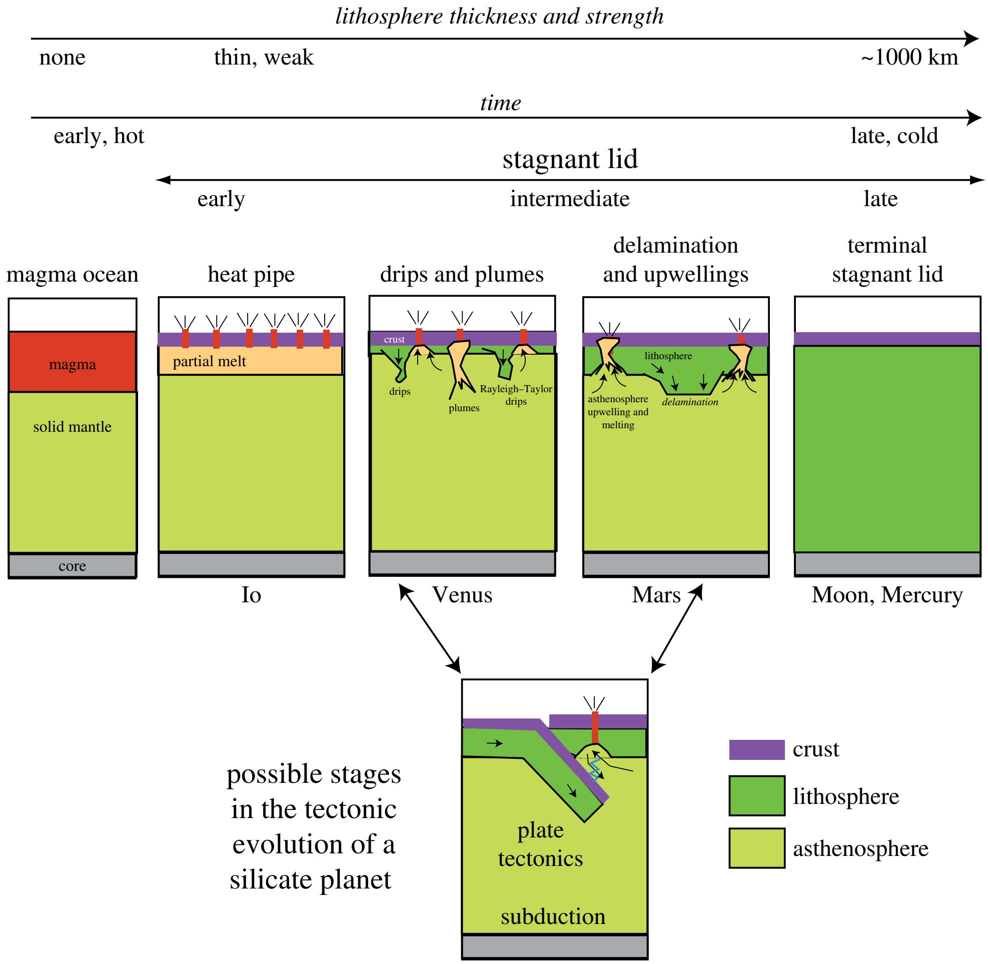

Figure 2. Possible evolution of tectonic styles for large silicate bodies like the Earth. Plate tectonics requires certain conditions of lithospheric density and strength to evolve and is likely to be presaged and followed by single lid tectonics. (Online version in colour.)

3. How did plate tectonics begin?

In order to understand how plate tectonics began, we need to understand how plates form and move today. By understanding the modern balance of driving and resisting forces, we can think more clearly about when this first existed such that a single lid could break to form plates and set these in motion; plate tectonics must have begun after that time. Geodynamic considerations compel the conclusion that plate motions today are mostly due to the sinking of oceanic lithosphere in subduction zones [16,17]. This is because modern oceanic lithosphere more than 30 Myr old is slightly (approx. 1%) denser than underlying asthenosphere [18] so old oceanic lithosphere will sink beneath lithosphere if possible. Such a density inversion was less likely in early Earth history because higher mantle potential temperatures were likely to be reflected in thinner mantle lithosphere (dense) and thicker mafic crust (buoyant). Given the existence of such a density inversion, the great strength of oceanic lithosphere is the most important impediment to starting and maintaining subduction zones (a global network of these and associated spreading ridges and transform faults is required for plate tectonics, not just one subduction zone). Today, oceanic lithosphere is generally strong but local zones of weakness exist that are significant enough to allow new subduction zones to form. A key observation is that of Gurnis *et al*. [19], who noted that about a third of all active subduction zones formed in the last 65 Ma, implying that subduction initiation must be a process that now occurs 'easily and frequently'. This process nevertheless requires a sufficiently long (>∼1000 km) lithospheric weak zone in order to nucleate a new subduction zone [20]. The great strength of undamaged oceanic lithosphere is demonstrated by the observation that no new subduction zones have nucleated along passive continental margins in the past 65 Ma (and longer) and that it is energetically more favourable for India to continue converging with Asia than for a new subduction zone to form to the south in the Indian Ocean [2]. Lithospheric strength at continent–ocean boundaries may have been lower in the past, in association with higher mantle potential temperatures and thinner lithosphere, as suggested by geodynamic modelling (e.g. [21]). A key point is that plate tectonics itself creates long weak zones in the oceanic lithosphere in the form of transform faults and their inactive equivalents, fracture zones, but such plate tectonic-induced lithospheric weaknesses would not have existed before plate tectonics began.

With this caveat in mind, the general sequence of events leading to the formation of a new subduction zone today is reasonably well understood. Sinking of old, dense oceanic lithosphere adjacent to a long (approx. 1000 km or more) transform fault and/or fracture zone is thought to have been where the present Izu–Bonin–Mariana (IBM) subduction zone nucleated. The IBM subduction zone formed when the western Pacific plate—which encompasses some of the oldest oceanic lithosphere on Earth [22]—began sinking approximately 52 Ma [23]. Initial subsidence of old oceanic lithosphere eventually began down-dip motion (true subduction) to torque the Pacific plate to move WNW and form the bend in the Emperor–Hawaii seamount/island chain approximately 47 Ma [24].

Long lithospheric weaknesses like transform faults and fracture zones are unlikely to have existed before plate tectonics began and are unlikely to have been exploited to form the first subduction zone. Other types of lithospheric weaknesses must have existed in order to form the first subduction zone and start plate tectonics. What processes might produce lithospheric weaknesses of sufficient length to start a new subduction zone?

Impacts of large bolides could rupture the lithosphere sufficiently to start a new subduction zone. Numerical modelling indicates that bolides with diameters greater than 500 km could damage the lithosphere sufficiently to start subduction [25]. Such large impacts are not likely after the first few hundred million years of Earth history and are thus unlikely to have been responsible for starting plate tectonics in Archaean or younger times, which is when most workers think is when plate tectonics started [26].

Another idea about how to develop lithospheric weaknesses in a single lid tectonic regime was offered by Bercovici & Ricard [27]. They suggested that lithosphere could be sufficiently damaged by a variety of non-plate tectonic processes such as mantle flow, transient subduction and dripping. Such processes promote shear localization leading to the establishment of long-lived weak zones that eventually—as Earth cools and oceanic lithosphere increases in thickness and density—collapse to form true tectonic plates driven by subduction alone.

It is also likely that the sufficiently large, hot and long-lived heads of starting mantle plums could rupture the lithosphere over a large enough region (greater than 1000 km diameter) to cause lithospheric collapse around the margins to create the first subduction zone needed to start plate tectonics [28]. Mantle plumes are ubiquitous features of active silicate bodies [3] and occurred on Earth all through geological time [29].

Water plays an essential role in localizing the lithospheric weaknesses that are needed to initiate and maintain subduction [30]. Water facilitates plate bending by penetrating down along bending-induced normal faults into the upper mantle, where it transforms strong, olivine-dominated peridotite into weak serpentinite [31], making it much easier to bend the plate. Water continues to be important for the sustainability of subduction because a weak zone composed of wet, weak sediments and serpentinites quickly forms on top of the sinking lithospheric slab. This weakness lubricates the plate interface and allows the plate to mostly slide down-dip (although approx. 10% of plate motion is vertical, resulting in trench rollback; [32]). Weakness above the subduction interface is enhanced due to the slab-released fluids and off-scraped sediments, allowing the subducted lithospheric slab to develop and maintain the asymmetric subduction zone configuration, which is required to pull the rest of the downgoing plate and drive plate motions [33]. Water is almost certainly necessary for plate tectonics to occur but the presence of water alone is not sufficient.

A final key point is that making the first subduction zone is necessary but not sufficient for starting plate tectonics. There must be large enough tracts of strong, gravitationally unstable lithosphere that can be induced to rupture, sink and move, and in this way evolve a global plate mosaic. Even when lithospheric conditions become suitable for plate tectonics, it is likely to take some time to accomplish the transition from single lid to global mosaic, a point that is explored below.

4. When did plate tectonics start?

In order to address this question, we must keep in mind several things. (i) The possibility that multiple episodes of stagnant lid and plate tectonics may have transpired in Earth history. (ii) When did the force balance needed to sustain plate tectonics first exist? (iii) Does continental crust formation require plate tectonics? (iv) What is the geological evidence and how do we interpret it? (v) Other phenomena explained by a Neoproterozoic start of plate tectonics. We consider these five topics separately below.

(a) The possibilities of multiple tectonic episodes

Discussing when plate tectonics began is often unnecessarily forced into a dichotomy: that today there is plate tectonics and at some time in the past this 'switched on' from some unspecified non-plate tectonic mode. The assumption that Earth has only experienced two tectonic modes (plate tectonics and pre-plate tectonics) impedes the full range of thinking about the problem that is required to solve it. Reducing the pre-plate tectonic mode to a single behavioural regime is inconsistent with what we know about other active silicate bodies in the Solar System, as discussed above, where three distinct styles of single lid behaviour are readily recognized: 'heat-pipe' single lid tectonics on Io, vigorous single lid tectonics on Venus and sluggish single lid tectonics on Mars (figures 1 and 2). Framing the question as a temporal dichotomy of pre-plate tectonics and plate tectonics can also lead to misidentifying significant changes in Earth products as signalling the start of plate tectonics, when the change could have been some transition in single lid behaviour. The danger of such misidentification is seen in discoveries of significant change in some Earth product at a given time and this change is interpreted to be when plate tectonics began. Examples of this 'assumed duality' approach include the appearance of eclogitic inclusions in diamonds at approximately 3.0–3.2 Ga [34], the change to more felsic continental crust as the 'great thermal divergence' inferred from chemical compositions of basalts approximately 2.5 Ga [15] and changes in Ni/Co and Cr/Zn ratios in Archaean terrigenous sedimentary rocks and Archaean igneous/metaigneous rocks interpreted as indicating the start of modern plate tectonics [35]. The changes that these studies recognize may just as well reflect a significant transition from one style of single lid behaviour to another, not the beginning of plate tectonics. Such logic traps are easily avoidable if the likelihood of multiple styles of single lid and multiple transitions from single lid to plate tectonics in Earth history is acknowledged.

There is increasing evidence that the Earth experienced multiple tectonic episodes. There is general acknowledgement that very early Earth experienced a magma ocean stage [36] and this cooled to form a protocrust. Earth's magma ocean may have evolved into a heat-pipe tectonomagmatic regime [37] whereby mafic and ultramafic lava flows without a mantle root were progressively erupted, buried by younger flows, and ultimately remelted to erupt again, as inferred for modern Io (figures 1 and 2). Heat-pipe or other post-magma ocean stages may have evolved into a single lid tectonic style, with growth of mantle lithosphere beneath the Hadean protocrust, and lithospheric thickening is likely to have continued into Archaean time.

Geoscientists are starting to identify Earth's single lid tectonic episodes: one is suggested for 2.8–2.9 Ga [12] and between 2.45 and 2.2 Ga [38]. The 'boring billion' between 1.8 and 0.8 Ga may have been another single lid episode.

The boring billion was characterized by remarkable environmental, evolutionary and lithospheric stability [39], with the formation of the Rodinia supercontinent near the end of this interval. Cawood & Hawkesworth [40] list seven characteristics of the boring billion: paucity of passive margins; absence of glacial deposits and iron formations; lack of significant seawater Sr isotope spikes; lack of phosphate deposits; high ocean salinity; abundant anorthosites and alkali granites; and limited orogenic gold deposits. Sr and C isotopic compositions of seawater-proxy carbonates change little during this time. The boring billion was also a long time when the complexity of life increased very slowly. These characteristics are very like that expected from a sluggish single lid episode.

(b) When did the force balance needed to sustain plate tectonics first exist?

Plate tectonics could not have begun on Earth until three conditions were satisfied: (i) large tracts of lithosphere became generally denser than underlying asthenosphere; (ii) large tracts of lithosphere became generally strong enough to remain intact in subduction zones and pull the attached surface plate; and (iii) lithosphere developed weak zones that were profound enough to rupture and become new plate interfaces. When in Earth history were these three conditions first satisfied? Let us consider lithospheric density first. The density of oceanic lithosphere (thermal lithosphere of [5]) is the sum of crustal and mantle contributions. Unless there is significant intraplate igneous activity, the crustal thickness is essentially established at the spreading ridge but the mantle contribution increases with age as the plate cools and mantle lithosphere thickens [18]. It is unlikely that oceanic lithosphere has always been as dense and as strong as it is today; specifically, it is likely to have been weaker and more buoyant in the past when the mantle was hotter. As noted in the previous section, the mantle has been cooling over Earth history. If we accept the inference that Earth's potential temperature was approximately 150–200°C higher 2.5 Ga ago than it is today [14,15], then the mantle lithosphere is likely to have been thinner than it is today [41]. Descent styles of thinner, weaker oceanic lithosphere have been explored quantitatively as a function of mantle potential temperature. Eclogitic dripping (Rayleigh–Taylor instabilities) is likely only at mantle potential temperatures (T_{p}) greater than 1600°C under a single lid tectonic regime [42]. Geodynamic modelling for $T_{\mathrm{p}} = 1425$–1600°C indicates that lithospheric mantle will separate from the crust and sink, a process that Chowdhury *et al*. [43] call 'peeling off'. Once separated from buoyant upper crust, lithospheric peels will break off and sink vertically into the mantle.

In addition to controlling lithospheric behaviour, higher asthenospheric potential temperatures would have led to more melting and generation of thicker oceanic crust with a given amount of decompression [44]. For the modern Earth, most of the strength of old oceanic lithosphere resides in the mantle, especially for old (greater than 30 Ma) seafloor [45]. Thinner mantle lithosphere and thicker oceanic crust expected for the Archaean would have resulted in oceanic lithosphere that was less likely to sink though underlying asthenosphere and hold together as it sank [42].

The last statement is increasingly supported by geodynamic modelling. There is not space here to summarize all of the significant contributions on this topic but a general sense of this rapidly advancing field is useful for understanding when plate tectonics was likely to have begun. Numerical modellers are increasingly attracted to the problem of early Earth tectonics, especially Archaean tectonics. Their largely independent and progressively more realistic experiments increasingly agree that plate tectonics (as described in the Introduction) could not have occurred given the likely strength and density of oceanic lithosphere of those times. This does not mean that subduction did not happen in the Archaean, only that weak, unstable subduction zones were unlikely to persist long enough to 'infect' the rest of the lithosphere and generate a global plate mosaic. Lithospheric buoyancy considerations were emphasized by van Thienen *et al*. [46] in their critique of Archaean plate tectonics. Moyen & van Hunen [47] combined geochemical data and geodynamic models to infer that a short-term episodic style of subduction was a viable style of early Earth tectonics. Their modelling results show how the low strength of slabs in a hotter Earth led to frequent slab break-off events that would have prevented modern-style, long-lived subduction systems, resulting instead in frequent cessation and re-initiation of subduction on a time scale of a few million years. Thébaut & Rey [48] used realistic rheological and thermal parameters to conduct coupled thermo-mechanical numerical experiments to show that the dominant Archaean tectonic regime was a vigorous single lid style called 'sagduction' in response to density inversions and thermal weakening due to the emplacement of thick continental flood basalts onto thin, hot protocontinental crust. Sizova *et al*. [21] conducted two-dimensional petrological–thermomechanical tectonomagmatic numerical experiments developed for conditions appropriate to the hotter early Archaean lithosphere and used the results to infer delamination and dripping of the lower primitive basaltic crust into the mantle, local thickening of the primitive basaltic crust and small-scale crustal overturn. They envisioned a vigorous single lid tectonic regime that was interrupted by short-lived subduction-like episodes. These results increasingly impel us to consider that early Earth experienced multiple episodes and styles of single lid behaviour.

(c) Continental crust formation: is it a plate tectonic signal?

Some geoscientists assume that, because continental crust is generated by plate tectonics today (mostly above subduction zones at convergent plate boundaries), the existence of continental crust of any age implies that plate tectonic processes must have operated in order to form such crust. This is a logical fallacy, because if this assumption was correct then the existence of continental crust as old as 3.8 Ga must indicate that plate tectonics began at least that long ago. In fact, most geoscientists agree that plate tectonics began significantly after 3.8 Ga [26], so the implicit linkage between formation of continental crust and plate tectonics should be critically considered.

There is no empirical or experimental evidence indicating that plate tectonics is required in order to generate continental crust. Formation of thickened crust of more felsic composition than expected for mantle melts (basalt) only requires hydrous magmatic conditions, including hydrous mantle melting and/or remelting of hydrous mafic crust. The importance of water for generating felsic melts has been known for decades. Campbell & Taylor [49] were among the first to argue that involvement of water in magmagenesis is required to make felsic continental crust. Hydrous melting of mantle tends to generate more Si-rich magmas (e.g. [50]), including high-Mg andesite and boninite, and such compositions can fractionate to form proportionately much larger volumes of felsic magma than can dry basaltic magmas. In addition, mafic rocks metamorphosed under hydrous conditions are likely to form amphibolite in the middle crust. Silica-rich magmas are expected from partial melting of amphibolite in the crust ($P \sim 1$ GPa) at $T \sim 750$–1000°C (e.g. [51,52]), conditions that were common in Archaean time. Subduction is an excellent way to introduce water into the mantle to trigger hydrous melting and to generate and melt amphibolites today but it is not the only way. All that is needed is that there be abundant surface water and a way to move this deep into the crust and upper mantle where it can participate in partial melting.

We know that surface water has been present on Earth since at least 4.3 Ga [53]. Amphibole will form wherever water interacts with mafic igneous rocks at $T \sim 400$–600°C; such conditions are characteristic of lower oceanic crust (e.g. [54]) and may reflect the root zones of seafloor hydrothermal vents, which were probably very abundant in Archaean times, as demonstrated by the abundant volcanogenic massive sulfide deposits of that age. Once water is fixed in mid-lower crustal amphibolite, this can be melted *in situ* to generate felsic magmas or can sink into the mantle via single lid processes like delamination or sagduction, where the water will be released and delivered to the surrounding mantle. A cartoon illustrating these processes is shown in figure 2. The point to be emphasized is that fractionation of hydrous mantle melts and/or partial melting of crust containing abundant amphibole is all that is needed to generate felsic melts and continental crust, and these processes do not require plate tectonics [7,42].

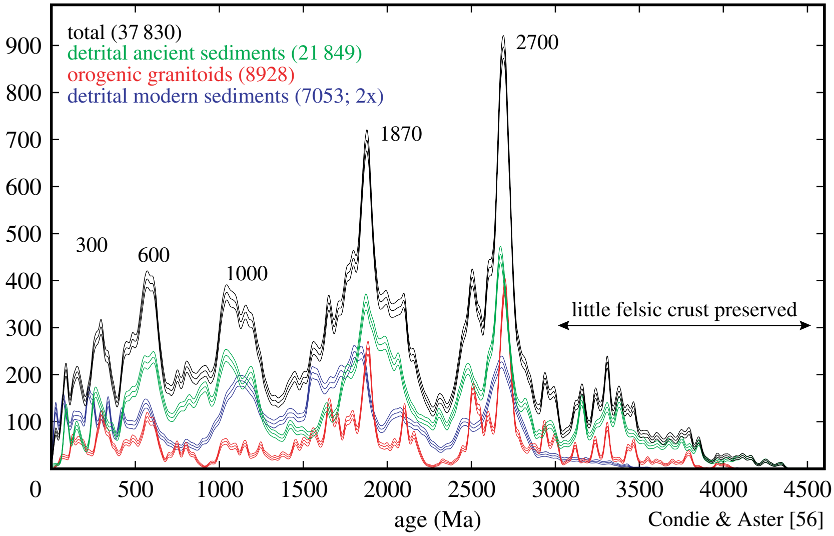

Figure 3. The detrital zircon age spectra, for orogenic granitoids, sedimentary rocks and modern river sediments, modified after Condie & Aster [56]. See text for further discussion.

Recent studies add detail to these observations. Moyen & Laurent [55] noted that Archaean granitic rocks are dominated by tonalites, trondhjemites and granodiorites (the TTG suite) formed by partial melting of alkali-rich metabasalt and amphibolite. Moyen & Laurent [55] acknowledged the arc-like trace element signatures of the TTG suite but cautioned that these signatures were controlled by the nature of the source and the conditions of melting, concluding that such controls and conditions are not directly linked to one particular tectonic setting. They noted that Archaean TTG suites were likely to be generated under a range of melting depths, from approximately 15 to 70 km deep (∼5 to >20 kbar). They emphasized that mafic crust could be partially melted to generate TTGs in a range of settings that did not require plate tectonics.

A slightly different perspective is provided by Rozel *et al.* [7]. These workers built on the understanding that hotter Archaean mantle temperatures resulted in two distinct styles of single lid behaviour: a heat-pipe regime dominated by volcanism (Io-like single lid) and the 'plutonic squishy lid' tectonomagmatic regime dominated by intrusive magmatism (Venus-like single lid). Rozel *et al.* [7] carried out numerical modelling of thermochemical convection to show that both regimes would be capable of producing TTG suites and thus continental crust. These numerical models showed that the volcanism-dominated heat-pipe tectonic regime was not able to produce felsic continental crust whereas the plutonic squishy lid tectonic regime dominated by intrusive magmatism was associated with hotter crustal geotherms and generated a variety of TTG magmas. Rozel *et al.* [7] concluded further that the pre-plate tectonic Archaean Earth operated in the plutonic squishy lid regime rather than in an Io-like heat-pipe regime, which may have been more important in the Hadean.

Geological evidence from the detrital zircon record (figure 3) strongly suggests that Eo-Archaean and Hadean crust was largely mafic [57]. Chowdhury *et al.* [43] addressed the dearth of such old crust using thermomechanical modelling. They inferred that such remnants were destroyed when continental crust began to stabilize in Late Archaean and Early Palaeoproterozoic time (3–2 Ga). Their numerical models indicate that non-plate tectonic recycling via lower crustal 'peeling-off' (delamination) was common during this interval and further inferred that destruction of the early mafic crust may have resulted in felsic magmatism that accelerated continental crust production by single lid magmatic processes.

One of the most important datasets that we have to constrain this discussion is the spectrum of zircon U-Pb ages. Zircon is generated when silicate magmas cool and is most abundantly produced by slow cooling of felsic magmas, such as TTGs or granite. Zircon age peaks are commonly interpreted either as crustal production peaks or as peaks of subduction-produced crust selectively preserved during continent–continent collision. The zircon age spectrum mostly reflects the proportion of zircon produced of a certain age, which reflects the amount of plutonic felsic rocks produced per unit time. There may have been quantitative loss of Hadean and Early Archaean zircons as this crust was mostly destroyed, but, once continental crust began to be preserved later in the Archaean, zircons of this age seems to have been largely preserved [58]. The record from about 3.0 Ga probably reflects crustal production, for the most part. Some felsic magmas produce more zircon per unit of magma than others (zircon fertility factor (ZFF) of [59]) but all felsic plutonic igneous rocks crystallize zircon and yield considerable zircon when they are eroded. The zircon age spectra for igneous rocks, sedimentary rocks of various ages and modern river sediments show similar trends, leading to confidence that these records contain useful information about the rate of felsic magma production and thus continental crust through time.

The zircon age spectra show two dominant peaks at 2700 and 1870 Ma (figure 3). The Late Archaean peak in the compilation of Condie & Aster [56] is somewhat larger than the mid-Palaeoproterozoic peak but they are broadly similar. The next largest peak approximately 1000 Ma is far smaller; this peak may be larger than representative because this age granitoids have very large ZFF ~3.5. After this there are minor peaks at 600 and 300 Ma.

It is very difficult to explain the large 1870 and 2700 Ma peaks as having formed by plate tectonic processes, these peaks are far too large by comparison with zircon/crustal production rates of slightly older and younger ages. Plate tectonics creates and destroys crust at a relatively constant rate, with some variation over the course of the supercontinent cycle [60]. In the compilation of Condie & Aster [56], the approximately 600 Ma Greater Gondwanaland peak is substantial but the approximately 300 Ma Pangea peak is not especially large. The 1870 and 2700 Ma peaks represent tremendous increases over zircon/crustal production just before and just after the peaks. These huge differences are most consistent with non-plate tectonic causes, for example mantle overturn and planetary resurfacing during variants of very vigorous single lid behaviour.

(d) What is the geological evidence and how do we interpret it?

Here we briefly consider what the geological record says about when the modern episode of plate tectonics began, and we should recognize that some kinds of evidence are more compelling than others. Large-scale features like ophiolites and high-pressure metamorphic rocks should be emphasized over small-scale lines of evidence like trace element ratios in sediments and igneous rocks or inclusions in zircons and diamonds because changes in the latter may reflect changes in single lid behaviour, as outlined above, whereas the former are diagnostic of plate tectonic processes. Figure 4 summarizes the temporal distribution of lithologies formed by plate tectonic processes in Cenozoic time and that certainly formed by plate tectonic processes in the past. The most diagnostic rock assemblages are ophiolites (figure 4*a*), blueschist and related subduction zone metamorphic assemblages (figure 4*b*), and ultra-high-pressure (UHP) metamorphic terranes and the collision gemstone ruby (figure 4*c*). These diagnostic lithologies are concentrated in Neoproterozoic and younger times, although a few 2.1–1.8 Ga ophiolites are also documented. More data have been added to the compilation, and these new data continue to validate the original conclusion of Stern [63] based on these three diagnostic assemblages that the modern episode of plate tectonics began in Neoproterozoic time. Just as important is the absence of these lithologies from ~1.0 to ~1.8 Ga.

It must be acknowledged that the geological record is incomplete and becomes more incomplete with age. Crustal recycling and erosion remove evidence, but there are many well-preserved outcrops of Archaean and Proterozoic rocks. The best way to overcome preservation bias is to consider multiple lines of evidence that are affected differently by erosion. Ophiolites commonly occur as the uppermost units of nappe stacks in orogenic belts and are especially vulnerable to erosion; in contrast, metamorphic rocks like blueschists and UHP terranes are

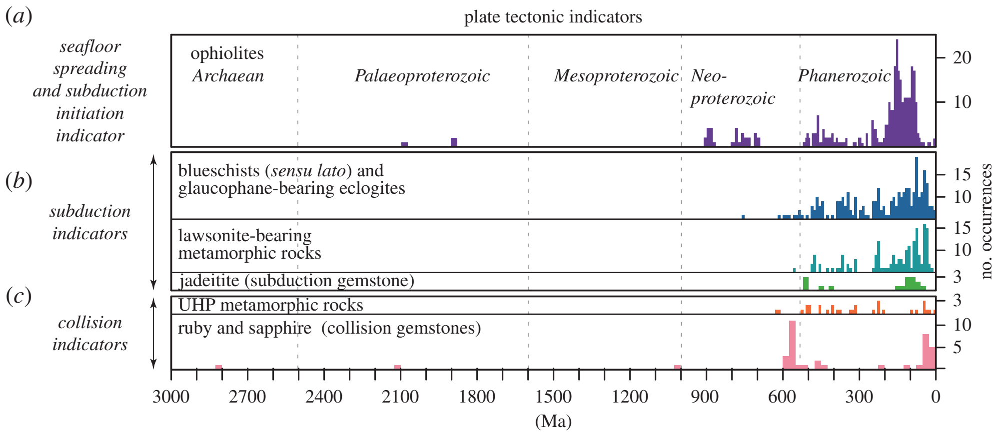

Figure 4. Comparisons of key petrotectonic indicators for plate tectonics and timings of major glacial episodes (vertical blue regions) spanning the last 3 Ga, modified from Stern & Miller [61]. (*a*) Ophiolites provide evidence of seafloor spreading, subduction initiation and horizontal motions consistent with plate tectonics (seafloor spreading and subduction initiation proxy). (*b*) Indicators of subduction zone metamorphism—blueschists, glaucophane-bearing eclogites, lawsonite-bearing metamorphic rocks and jadeitites—form only in the cool, fluid-rich environments in and above subduction zones (subduction proxies). (*c*) UHP metamorphic rocks and the gemstone ruby proxy continental collision and deep subduction of continental crust (continental collision proxies). Modified after Stern *et al*. [62]. (Online version in colour.)

uncovered by erosion and are difficult to obliterate. The fact that the multiple lines of evidence shown in figure 4 agree indicates that preservation bias is not a serious concern for the Archaean and Proterozoic record.

Palaeomagnetic evidence is potentially a key constraint but becomes less reliable with age [64]. Critical evaluation of this line of evidence for the purposes of understanding when palaeomagnetic constraints are robust and when they are not, and, on this basis, what the palaeomagnetic evidence says about when cratons moved relative to each other (=plate tectonic episodes) and when they did not (single lid), will require dedicated syntheses by members of the practising palaeomagnetic community.

The absence of ophiolites, blueschist and UHP terranes for approximately 800 Ma—approximately the time of the 'boring billion' or Earth's middle age [40]—provides further support that this was a single lid interval, as shown in figure 5. It is also significant that there are some 1.8–2.1 Ga ophiolites, suggesting that this may have been a short 'proto-plate tectonic episode' (figure 5). Nevertheless, it is increasingly acknowledged that Earth went through major tectonic changes in Neoproterozoic time (e.g. [65–67]). The appearance of low-T, high-P metamorphism indicative of 'cold subduction' begins approximately 800 Ma [68]. The evidence is overwhelming that this was when modern plate tectonics as defined at the start of this paper began.

(e) Other phenomena explained by a Neoproterozoic start of plate tectonics

Three additional Earth history mysteries are explained by recognizing that the modern episode of plate tectonics began in Neoproterozoic: the kimberlite record, Neoproterozoic snowball Earth, and the acceleration of biological evolution. These revelations are discussed briefly below.

Kimberlite eruptions—the water- and carbon dioxide-charged explosions from deep in the lithosphere that carry diamonds to the surface—are concentrated in Neoproterozoic and younger time (figure 6*a*). This has long been recognized but until recently unexplained. Stern *et al*. [62] argued that kimberlites became increasingly abundant from the Neoproterozoic onwards because subduction began to deliver increasingly large volumes of water and CO_2 deep into the mantle

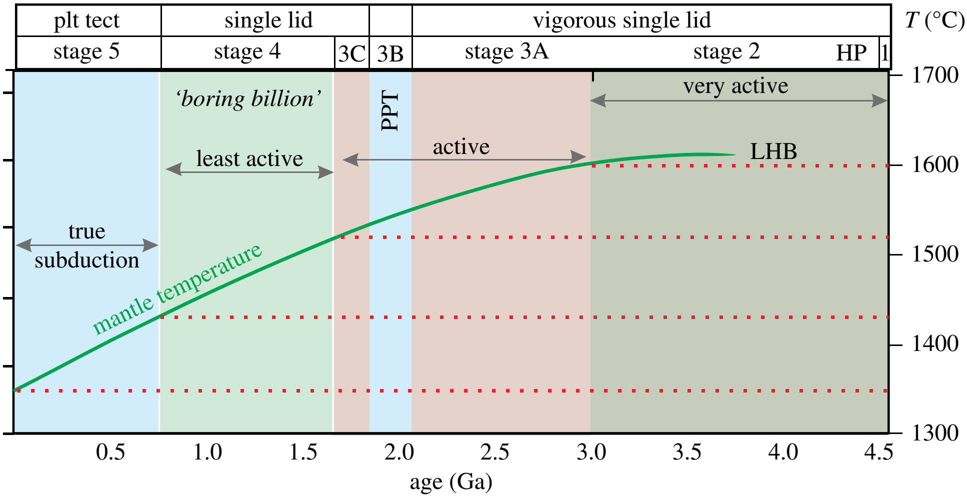

Figure 5. Modified version from Hawkesworth *et al*. [65] to reflect the author's preferred interpretation of Earth's tectonic history. HP, heat-pipe tectonics; LHB, Late Heavy Bombardment; PPT, proto-plate tectonics. (Online version in colour.)

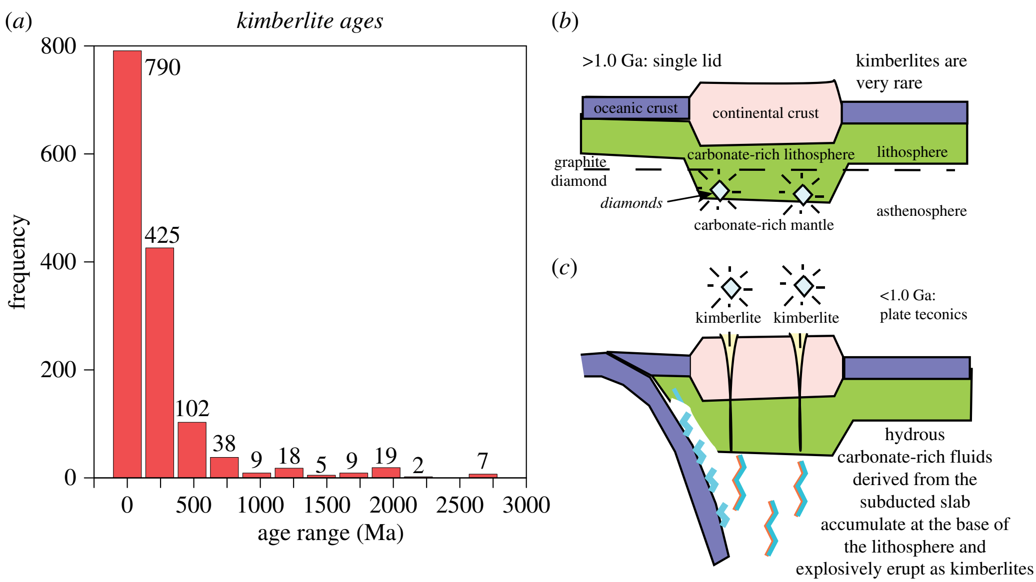

Figure 6. Kimberlites and the evolution of plate tectonics. (*a*) Histogram of kimberlite ages, binned each 500 Ma. Note the great increase in kimberlite in Neoproterozoic and younger time (less than 750 Ma). (*b*,*c*) Simple explanation why abundance of kimberlites increased approximately 750 Ma. (*b*) Before approximately 1 Ga, no plate tectonics and no deep subduction. Flux of water to mantle is low so fluid pressure at the top of the asthenosphere is low. (*c*) After approximately 1 Ga, plate tectonics and deep subduction delivers more water deeper into the mantle. Upward-infiltrating water interacts with carbonated peridotite mantle, generating abundant H_2O–CO_2 fluids and increasing fluid pressure at the top of the asthenosphere. Eventually build-up of fluid pressure breaks to the surface as kimberlite. Modified after Stern *et al*. [62]. (Online version in colour.)

then. Large quantities of water pumped deep into the mantle slowly percolated up from deeply subducted slabs to destabilize mantle carbonates. The combined H_2O–CO_2 fluid accumulated at the base of the lithosphere, resulting in especially high volatile pressures at the base of the cratonic lithosphere. Volatile accumulation continued until it was released by explosive eruptions from this depth to form kimberlites. Kimberlites did occur before Neoproterozoic time but they

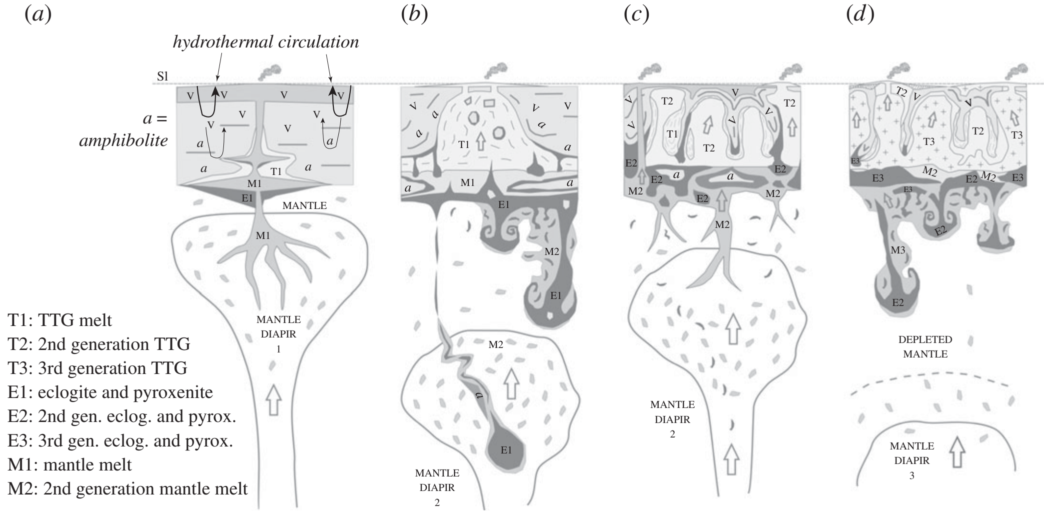

Figure 7. Cartoon modified after Bédard [69] showing how water might be delivered to the upper mantle and how felsic magmas (TTG suites) might be generated by melting crustal amphibolite. (*a*) A large mantle plume releases melt (M1) that constructs a thick volcanic crust, the lower part of which is metamorphosed into amphibolite. Underplating magma, which causes melting at the base of the crust, forms a first generation of tonalitic melt (T1) with complementary eclogitic to pyroxenitic restites (E1). (*b*) Smaller delaminated bodies mix into the shallow upper mantle and trigger the formation of a second generation of mantle melt (M2). (*c*) The first generation crustal restites are largely destroyed as M2 melts are generated, collect and ascend. New melt from a second mantle diapir also contributes to M2. M2 underplates the crust to form a second generation of tonalite melt (T2) by melting amphibolites, yielding a second generation of restites and cumulates (E2). Older tonalites (T1) are extensively remobilized at this time, and also contribute to T2. The voluminous T2 tonalites are buoyant and trigger a second cycle of partial crustal convective overturn. (*d*) As M2 magmatism wanes, the underplated layer cools and crystallizes. The restites and cumulates (E2) delaminate into the mantle, triggering the formation of a third generation of mantle melts (M3), and destroying the second generation restites. Melting of underplated M2 melt and relict lavas generate a third generation of tonalitic to granodioritic melt (T3), also yielding a third generation of restites and cumulates (E3). Older tonalitic rocks (T1 and T2) are extensively remobilized and represent the dominant part of T3. The voluminous T3 tonalites/granodiorites are buoyant and trigger a third cycle of partial convective overturn in the crust.

were much rarer than after Neoproterozoic time (figure 6*a*). Single lid tectonics could deliver small volumes of surface fluids and volatiles to the upper mantle by sagduction and drips (figures 7 and 6*b*) but the massively greater delivery of fluids and volatiles deep into the mantle needed to cause frequent kimberlite eruptions could not occur until plate tectonics and sustained subduction got underway in Neoproterozoic time (figure 6*c*). An alternate interpretation is offered by Tappe *et al*. [70].

Neoproterozoic snowball Earth is another mystery in Earth history that can be explained by a Neoproterozoic onset of the modern episode of plate tectonics. Stern & Miller [61] noted that the transition from Mesoproterozoic single lid tectonics to plate tectonics should have disturbed the climate equilibrium established by the previous billion-year-long stasis of silicate weathering–greenhouse gas feedbacks by re-distributing continents, increasing explosive arc volcanism, creating relief and stimulating mantle plumes. Formation of subduction zones could have redistributed mass sufficiently to caused true polar wander. These disruptions could have caused spectacular carbon isotope variations along with several episodes of Neoproterozoic snowball Earth. Stern & Miller [61] argued that the transition to plate tectonics could have caused nearly all of the proposed geodynamic and oceanographic triggers for Neoproterozoic snowball Earth events, and could also have contributed to biological triggers. Only extraterrestrial triggers cannot be reconciled with the hypothesis that the Neoproterozoic climate crisis was caused by a transition from single lid to plate tectonics.

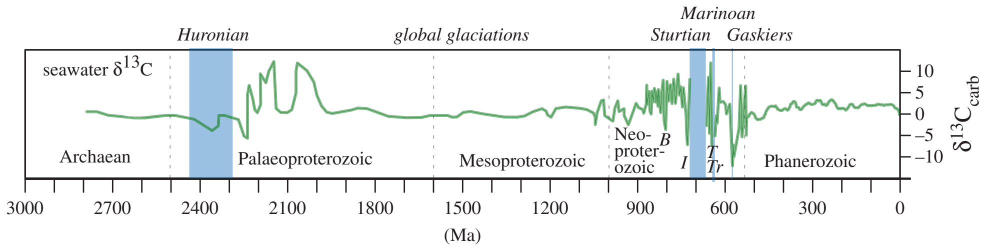

Figure 8. Timings of major global glacial episodes (vertical blue regions) spanning the last 3 Ga superimposed on a seawater δ^{13}C curve. Neoproterozoic C-isotope excursions B, I, T, Tr and S denote Bitter Springs, Islay, Tayshir, Trezona and Shuram anomalies, respectively. Note that the Neoproterozoic was a remarkable episode of change on Earth's surface, consistent with a transition from single lid tectonics to plate tectonics during Neoproterozoic time. Modified from [61]. (Online version in colour.)

An intriguing result of the suggestion that the transition from single lid to plate tectonics caused the Neoproterozoic climate crisis is that the stratigraphic record provides information for how long it took for the transition from single lid to formation of the first subduction zone and a two-plate planet to development of a global plate mosaic to occur. If we assume that individual 'snowball' episodes (Sturtian, Marinoan and Gaskier; vertical blue lines in figure 8) and major C isotope excursions (Bitter Springs, Islay, Trezona, and Shuram; figure 8) reflect the formation of new plates and subduction zones, then it took about 230 Ma for a global plate mosaic to form. There may not be a simple one-to-one correspondence between these climate and isotopic excursions on the one hand and formation of new plates and subduction zones on the other, but the approximately 230 Ma interval of climate and isotopic instability may approximate how long it took to accomplish the transition from the climate and isotopic equilibrium associated with the Mesoproterozoic single lid tectonic regime to that of the modern plate tectonic episode.

Finally, the start of plate tectonics in Neoproterozoic time provides new insights into the question of why biological evolution accelerated at that time. Eukaryotic life experienced a major diversification approximately 800 Ma [71]. The Neoproterozoic also witnessed the first appearance of marine planktonic single cellular nitrogen-fixing cyanobacteria and non-nitrogen-fixing picocyanobacteria [72]. The oldest known animal body fossils are found in approximately 571–566 Ma sediments. Trace fossil evidence of bilateral locomotion also occurs in *ca* 565 Ma sediments. Simple skeletonized metazoans occur only in rocks deposited during the last 8–9 million years of the Ediacaran Period [73]. Metazoans began to take over ecosystems during the Ediacaran (635 to 541 Ma) but the developmental toolkits were established in the Cryogenian (720 to 635 Ma) [74]. Biologists are very interested to know what stimulated this burst of diversification.

The link between Neoproterozoic tectonics and biological evolution is acknowledged [75], but a link between acceleration of biological evolution rates and the start of plate tectonics has not been considered heretofore. Such a link makes sense because creation of new habitats, isolation and interspecies competition is difficult and rare for single lid tectonic regimes but is easy and common with continents and plate tectonics (figure 9*b*–*g*). It is perfectly consistent with interpreting a single lid regime for the boring billion (*ca* 1800 to 800 Ma) that biological evolution proceeded so slowly during this interval. In contrast, new habitats are constantly being created and destroyed by plate tectonics. New continental margins form during continental break-up, increasing habitats and isolation, which spur diversification and speciation. These habitats are destroyed and species occupying similar niches are forced to compete when continents collide. By rapidly creating and destroying new habitats, plate tectonics expedites biological evolution [76]. Single lid tectonic regimes cannot stimulate biological evolution, although it is difficult to quantify the effects of these different evolutionary determinants. Nonetheless, a transition from

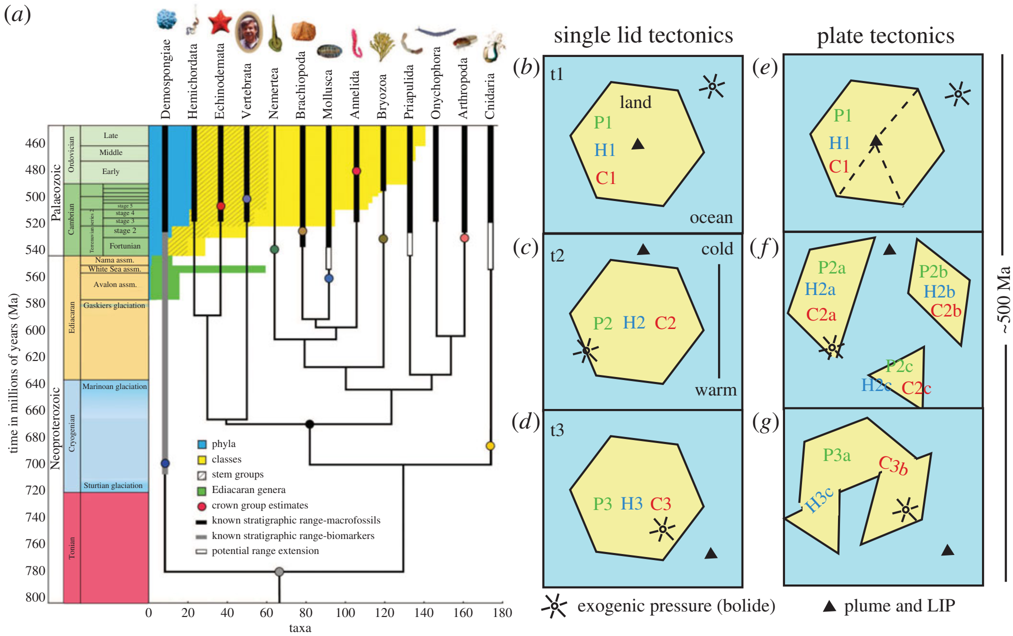

Figure 9. Animal evolution in the context of a Neoproterozoic transition to plate tectonics from Mesoproterozoic single lid tectonics. (*a*) Animal fossil record compared with the molecular divergence estimates for 13 different animal lineages. The known fossil record of animals is shown at the class and phylum levels (hatching indicates 'stem' lineages, i.e. lineages that belong to a specific phylum but not to any of its living classes); green shows the record of macroscopic Ediacaran fossils (see scale at bottom). Shown in thick black lines are known fossil records of each of these 13 lineages through the Cryogenian–Ordovician; most lineages first appear in Cambrian time, consistent with the animal fossil record. The extent of these stratigraphic ranges mirrors molecular estimates for each crown group (coloured circles), highlighting the general accuracy of the molecular clock [74]. (*b*–*g*) Cartoons showing how natural selection and evolution vary on a simplified Earth-like planet with subequal areas of continents and oceans and three interdependent life forms (plant 'P', herbivore 'H' and carnivore 'C') over a supercontinent cycle at three different times (1, 2 and 3) at approximately 100 million year intervals. Top of each panel corresponds to high latitudes (cool, arctic), bottom is equatorial (warm). Exogenic evolutionary pressures exist regardless of whether or not plate tectonics occurs, as does mantle plume activity including Large Igneous Provinces (LIPs). (*b*–*d*) Planet without plate tectonics, little change in continental configuration, only climatic isolation, few barriers, and slow climate change. Evolutionary pressures are dominantly biological and exogenic. (*e*–*g*) The situation for the same planet with plate tectonics over the course of a supercontinent cycle. This provides many opportunities for isolation, diversification under different conditions of natural selection and evolution; this is 'rift pump'. Evolutionary rift pumping continues until continents collide, when different species co-mingle and compete and new ecological systems are established; this is 'collision pump'. Endogenic evolutionary pressures are more important and are dominated by plate tectonics effects.

Mesoproterozoic single lid to Ediacaran plate tectonics helps explain the accelerating pace of biological evolution beginning in Neoproterozoic time.

5. Conclusion

Below are 10 observations resulting from considerations explored in this paper:

1. Plate tectonics is a very unusual convective style for a silicate planet. All other active silicate bodies are encased in a single lithospheric lid.
2. What is common in space may also be common over Earth history, and a significant part of Earth history likely involved single lid tectonic regimes. Plate tectonic and single lid tectonic regimes may have alternated over this history.

3. Geodynamic studies conclude that it is the sinking of dense oceanic lithosphere in subduction zones that drives plate motions. Because Earth's upper mantle has cooled 150–200°C since Archaean time, dense oceanic lithosphere has thickened and buoyant oceanic crust has thinned, making the lithosphere denser—and plate tectonics more likely—with time.
4. There is a wide range of single lid tectonic regimes observed on Io (heat-pipe single lid), Venus (vigorous single lid) and Mars (sluggish single lid). These variations in single lid behaviour largely reflect the sub-lithospheric thermal structure of the silicate body and lithospheric thickness, strength and composition. A variety of single lid styles is also expected to have characterized Earth as its mantle cooled and lithospheric mantle thickened and oceanic crust thinned. Possible variants include heat-pipe, plutonic squishy lid, stagnant lid and sluggish lid, and these single lid modes are variously affected by sagduction, eclogitic dripping, delamination and plume–lid interactions. Earth's tectonic regime during the boring billion/Earth's Middle Age (approx. 1.7–0.8 Ga) was probably a protracted episode of sluggish single lid.
5. The transition from single lid to plate tectonics requires a lithospheric weakness approximately 1000 km long or longer. Early in Earth history this could have formed by large bolide impacts. Later in Earth history progressively damaged lithosphere and/or plumehead–lithosphere interactions would have provided the weak zones needed for the first subduction zone to form. Subsequent subduction zones needed to make the global plate mosaic may have formed similarly or exploited new weakness caused by plate tectonics such as transform faults and fracture zones.
6. The formation of continental crust and plate tectonics are distinct processes and should not be conflated because continental crust can and almost certainly did form without plate tectonics. The zircon U-Pb age spectra need to be critically reconsidered in light of the likelihood of single lid tectonic episodes in Earth history. The scarcity of zircons greater than 3.0 Ga indicates that little felsic continental crust was produced before this time and/or preserved, implying a thin, weak lithosphere dominated by basaltic and komatiitic crust that easily collapsed back into the mantle. Plate tectonics produces and destroys continental crust at a quasi-continuous rate, so large peaks at approximately 2.7 Ga and approximately 1.9 Ga are unlikely to have formed by plate tectonic processes. The dominant peak approximately 2.7 Ga may mark a major episode of mantle overturn leading to widespread granite melting, while the approximately 1.9 Ga peak may mark a similar episode of mantle overturn associated with proto-plate tectonics, inferred from the presence of ophiolites about this time. The smaller variations in U-Pb zircon ages observed for the past approximately 1 Ga are more consistent with modern plate tectonic processes than the larger variations observed for older rocks.
7. Plate tectonics produces unique igneous assemblages such as ophiolites as well as unique metamorphic assemblages such as blueschists and UHP terranes. With the exception of a brief episode approximately 2.0 Ga when some ophiolites formed, these assemblages do not appear in the rock record until Neoproterozoic time, signifying that the modern episode of plate tectonics began in the past 1 Ga.
8. The increase in the abundance of kimberlites since the Neoproterozoic is explainable as being due to the onset of massive transport of surface water and CO_2 deep into the Earth (mantle ingassing) due to subduction, which began on a large scale in the Neoproterozoic. These volatiles migrate upward from deeply subducted slabs and are trapped at the base of cratonic lithosphere. Build-up of volatile pressure at the base of cratonic lithosphere over many millions of years leads to explosive eruption of kimberlites.
9. The transition from the boring billion single lid tectonic regime to plate tectonics should have disrupted Earth's climate equilibrium. An otherwise inexplicable major episode of dramatic climate and C isotope swings known as Neoproterozoic snowball Earth was likely to have been caused by this transition.

10. The duration of Neoproterozoic climate and isotopic variations suggests that the transformation from single lid to the first subduction zone (two plates) to a complete global, plate tectonic mosaic took approximately 230 Ma to accomplish.

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